

Holocene carbon cycle dynamics

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We are investigating the late Holocene rise in CO₂ by performing four experiments with the climate-carbon-cycle model CLIMBER2-LPJ. Apart from the deep sea sediments, important carbon cycle processes considered are carbon uptake or release by the vegetation, carbon uptake by peatlands, and CO₂ release due to shallow water sedimentation of CaCO₃. Ice core data of atmospheric CO₂ between 8 ka BP and preindustrial climate can only be reproduced if CO₂ outgassing due to shallow water sedimentation of CaCO₃ is considered. In this case the model displays an increase of nearly 20 ppmv CO₂ between 8 ka BP and present day. Model configurations that do not contain this forcing show a slight decrease in atmospheric CO₂. We can therefore explain the late Holocene rise in CO₂ by invoking natural forcing factors only, and anthropogenic forcing is not required to understand preindustrial CO₂ dynamics.

1. Introduction

During the last 8000 years (8 ka) the Antarctic ice core records [*Inderm Ehle et al.*, 1999; *Monnin et al.*, 2004] show an increase in atmospheric CO₂ by about 20 ppmv that has so far proven difficult to explain. Possible explanations of the rise in atmospheric CO₂ include a decrease in terrestrial carbon storage [*Inderm Ehle et al.*, 1999; *Brovkin et al.*, 2002], increases in sea surface temperatures [*Inderm Ehle et al.*, 1999; *Joos et al.*, 2004], coral reef regrowth [*Ridgwell et al.*, 2003], and carbonate compensation [*Broecker et al.*, 1999; *Elsig et al.*, 2009]. *Broecker et al.* [1999] suggested that reconstructed reduction in the carbonate ion content of the deep sea was caused by carbonate compensation to the early Holocene forest regrowth. *Elsig et al.* [2009] estimated that 15 ppmv growth of CO₂ during the Holocene could be explained by carbonate compensation to a land biosphere uptake of 700 and 200 GtC prior and during the Holocene, respectively. The remaining CO₂ increase they attributed to coral reef growth and other mechanisms. Last but not least, *Ruddiman* [2003] proposed that the CO₂ growth was caused by human activities including CO₂ emissions due to slash and burn agriculture.

The magnitude of emissions necessary to increase CO₂ by 20 ppmv (ca. 200 GtC) due to agricultural changes is not supported by estimates of landuse emissions based on population estimates prior to 1700 AD [Joos et al., 2004; Pongratz et al., 2009], though assuming a fixed relation between population and agricultural area may severely underestimate deforestation [Kaplan et al.,

2009]. Besides, an increase in land surface albedo, offsets the warming due to the CO₂ increase. Explaining the rise in CO₂ by a decrease in natural land carbon storage also appears unlikely because the terrestrial biosphere likely takes up more CO₂ under elevated CO₂ levels due to CO₂ fertilization [Kaplan et al., 2002; Joos et al., 2004]. Besides, substantial amounts of peat have accumulated in boreal wetlands during the Holocene [Gajewski et al., 2001]. Therefore it is unlikely that land processes are an important contributor to CO₂ growth.

In this paper we test the explanations for the Holocene rise in atmospheric CO₂ based on changes in the oceanic carbonate chemistry applied before the background of the other natural mechanisms, including vegetation dynamics, and peat accumulation.

2. Model description and experiments

2.1. CLIMBER2-LPJ

To investigate the dynamics of Holocene CO₂ we are using CLIMBER2-LPJ in four experiments. CLIMBER2-LPJ consists of the earth system model of intermediate complexity (EMIC) CLIMBER2, coupled to the dynamic global vegetation model (DGVM) LPJ.

CLIMBER2 [Petoukhov et al., 2000] consists of a 2.5-dimensional statistical-dynamical atmosphere with a resolution of roughly 51° (longitude) by 10° (latitude), a zonally averaged ocean resolving three basins with a latitudinal resolution of 2.5°, and a sea ice model. CLIMBER2 also contains oceanic biogeochemistry, a model for marine biota, and a sediment model [Archer, 1996; Brovkin et al., 2002, 2007]. Weathering rates scale to runoff from the land surface. In comparison to the earlier study by Brovkin et al. [2002], this setup allows to simulate carbonate compensation interactively. This is also an advantage compared to Elsig et al. [2009] who used a box diffusion model of the ocean uptake with a simple parameterization of carbonate compensation.

To this EMIC we have coupled the DGVM LPJ [Sitch et al., 2003] in order to investigate land surface processes at a significantly higher resolution of 0.5x0.5°. We also implemented carbon isotope fractionation according to Scholze et al. [2003]. Monthly anomalies from the climatology of the climate fields are passed to LPJ, where they are added to climate patterns based on the Climatic Research Unit CRU-TS climate data set [New et al., 2000]. The carbon flux F_{AL} between atmosphere and land surface is determined from the annual change in the LPJ carbon pools, and employed in CLIMBER2 to determine the CO₂ concentration. Biogeochemical feedbacks are thus determined by the combination of CLIMBER2 and LPJ, while biogeophysical effects are solely determined by CLIMBER2.

2.2. Model initialization and experiments

On multi-millennial timescales the global carbon cycle is never in a complete equilibrium state due to small but persistent fluxes associated with weathering (e.g. Munhoven, 2002) paced by climate changes during glacial cycles. To get consistent non-equilibrium initial conditions, the simulations should start long before the early Holocene, preferably at the last glacial inception. Such a non-equilibrium approach is still beyond the computational efficiency of 2- or 3-dimensional Earth system models.

To get the initial state of the carbonate system for 8 ka BP, the starting point of our experiments, we used the following two-step initialization approach:

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1. The model was run with present-day equilibrium conditions with CO_2 at 280 ppmv. This simulation assumes carbonate sedimentation of 9 Tmol/a CaCO_3 in the deep ocean [Brovkin *et al.*, 2007] and 8.4 Tmol/a in the shelf areas. Volcanic outgassing is equal to 5.85 Tmol/a CO_2 , added to the atmosphere, which is a steady state for the coupled system under present-day conditions. Using this setup, the model was run for 20,000 years.

2. Using the model state reached in step 1, we changed the orbital configuration to 8 ka BP and prescribed CO_2 to 260 ppmv. Ocean alkalinity was increased to get a carbonate sedimentation flux of 16 Tmol/a in the deep ocean and 2 Tmol/a on the shelves in order to simulate the maximum in CaCO_3 preservation in the deep sea before the onset of the Holocene, clearly seen in the deep sea CaCO_3 records [Broecker *et al.*, 1999]. Atmospheric $\delta^{13}\text{CO}_2$ was initialized at -6.4 ‰, as measured in ice cores for 8 ka BP [Elsig *et al.*, 2009]. The model was run with prescribed CO_2 for 5000 years to insure that the CO_2 distribution in the ocean stays in equilibrium, and finally for another 2000 years with CO_2 and $\delta^{13}\text{CO}_2$ exchanged interactively.

This simulation setup ensures that the system is in equilibrium with the initial conditions, a mixture of weathering conditions for present-day and the Holocene CaCO_3 preservation maximum. LPJ was then spun up separately for 2000 years under 8 ka BP boundary conditions. Remains of the Laurentide ice sheet at 8 ka BP are neglected.

In the transient simulations the climate model was driven by orbital forcing with interactive CO_2 and $\delta^{13}\text{CO}_2$. The specific forcings for the carbon cycle were:

1. Shallow water sedimentation (SWS). We prescribed a constant accumulation of carbonates on the tropical shelves of 15.5 Tmol/a, produced by coral reef growth and general accumulation on the shelves, for example through shelf bioherms. The magnitude of this forcing is in line with the estimate by Milliman [1993] and the model value by Ridgwell *et al.* [2003], but higher by about 5 Tmol/a than the estimate used by Kleypas [1997].

2. Peat accumulation forcing. Gajewski *et al.* [2001] estimate that peat deposits of about 450 GtC have accumulated over the last 21,000 years, though other estimates are substantially lower, e.g. 273 GtC estimated by Turunen *et al.* [2002]. The main reason for this difference is that Gajewski *et al.* use a higher estimate for the bulk density of peat. We prescribe peat accumulation with an accumulation of 100 GtC in 8 ka, about one third of the Gajewski *et al.* [2001] estimate for the last 8 ka.

Table 1. Carbon cycle components considered in the experiments.

Experiment	Vegetation	Peat	Shallow water sediment
AO	-	-	-
AOV	+	-	-
AOVP	+	+	-
AOVPC	+	+	+

We performed the four model experiments shown in Table 1. In experiment AO (atmosphere-ocean), the atmospheric CO_2 concentration is determined solely by the marine carbon cycle in CLIMBER2, C fluxes F_{AL} between atmosphere and land surface are neglected. In AOV (atmosphere-ocean-vegetation) the model setup is as in AO, but fluxes F_{AL} are considered. In experiment AOVP (atmosphere-ocean-vegetation-peat), peat accumulation is considered in addition to setup AOV. Finally, in experiment AOVPC (atmosphere-ocean-vegetation-peat-corals) we are also considering CO_2 outgassing due to shallow water sedimentation of CaCO_3 .

3. Results

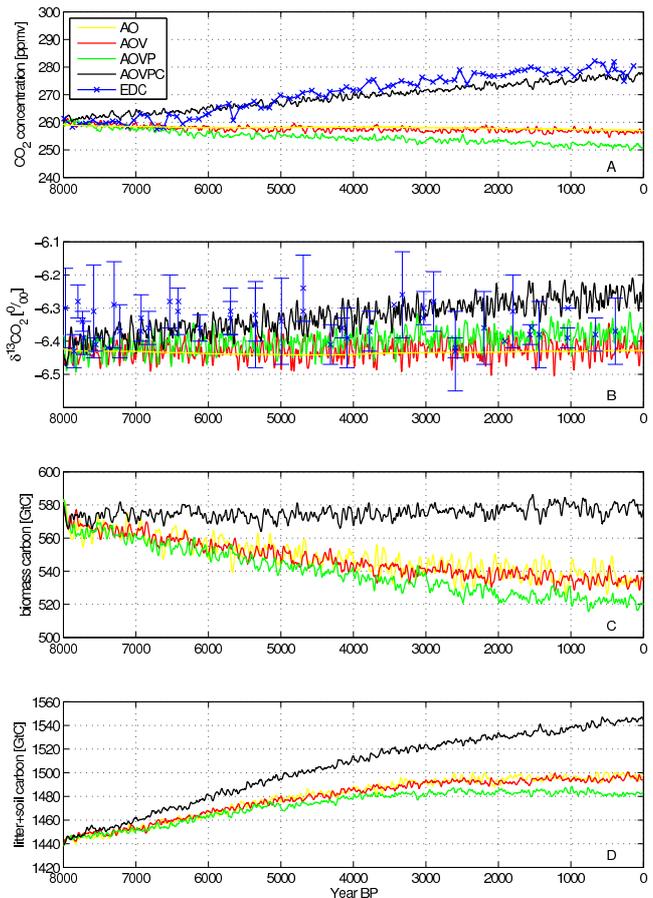


Figure 1. CO_2 concentration [ppmv] (A) and $\delta^{13}\text{CO}_2$ [‰] (B) in the model experiments, as well as data from EPICA Dome C [Monnin *et al.*, 2004; Elsig *et al.*, 2009]. Dynamics of terrestrial biomass [GtC] (C), and litter and soil carbon [GtC] (D) in the experiments. Model data are smoothed with a 50 year Gaussian weighted filter.

Fig. 1 (A) shows the atmospheric CO_2 concentration for the four experiments. In AO, CO_2 decreases from 259 ppmv at 8 ka BP to 257 ppmv at present day. This is the result of (i) a small drift in the carbonate system tuned to the present-day carbonate state, and (ii) the direct impact of the changes in orbital forcing between 8 ka BP and now, a (very small) uptake of carbon by the ocean. If biomass changes are considered in AOV there is no substantial change from AO.

In AOVP, where peat uptake is considered, the decrease in CO_2 is slightly stronger: CO_2 decreases from 259 ppmv to 251 ppmv, and down to 243 ppmv in a sensitivity experiment with doubled uptake (not shown). Finally, in AOVPC, shown in black, where SWS is considered, CO_2 increases from 259 ppmv to about 278 ppmv due to the decreasing total alkalinity in response to CaCO_3 sedimentation. Comparing these experiments to the EPICA Dome C data [Monnin *et al.*, 2004], it becomes clear that experiment AOVPC matches measurement data closely, while no other experiment can reproduce the CO_2 increase.

Fig. 1 (B) shows $\delta^{13}\text{C}$ of atmospheric CO_2 in our experiments. $\delta^{13}\text{C}$ allows conclusions with regard to the partitioning of carbon between the biosphere and other carbon pools, since plants preferentially assimilate C^{12} as opposed to C^{13} . In AO and AOV,

$\delta^{13}\text{C}$ stays constant, as expected if the land biosphere doesn't take up carbon. In AOV, $\delta^{13}\text{C}$ increases slightly, and in AOVPC it increases more strongly, the latter being a result of the increase in biomass through CO_2 fertilization. The spline fit through the EPICA Dome C record of $\delta^{13}\text{C}$ by *Elsig et al.* [2009] shows a strong increase in $\delta^{13}\text{C}$ between 11 ka BP and 6 ka BP, after which $\delta^{13}\text{C}$ slowly decreases. The error bars of the raw measurement data allow other temporal evolutions as well, including a slow increase in $\delta^{13}\text{C}$ between 8 ka BP and about 2.5 ka BP. Our results for experiment AOVPC are within the error bars of the *Elsig et al.* [2009] $\delta^{13}\text{C}$ record, though they leave this range at about 2.5 ka BP. This divergence is no surprise, since anthropogenic deforestation is not considered in our experiments, which would certainly lead to a decrease in $\delta^{13}\text{C}$.

Total biomass carbon, shown in Fig. 1 (C), decreases in all experiments but AOVPC due to the decrease in high latitude insolation and the corresponding reduction of boreal forest. In AOVPC there is a slight increase in biomass. Similarly, the litter and soil carbon (D) increases in all experiments, with experiment AOVPC showing the largest increase, about 105 GtC. Experiments AO, AOV, and AOVPC have very similar effects on modelled land vegetation: Biomass decreases, and the carbon stored in vegetation at 8 ka BP is shifted into litter and soil, making the land biosphere carbon neutral on the timescales considered.

In experiment AOVPC the biomass increases due to CO_2 fertilization, and there is a corresponding increase in both litter and soil carbon, i.e., the land surface is a carbon sink. This is in disagreement with *Brovkin et al.* [2002], who show a decrease of about 90 GtC in terrestrial carbon storage in the VECODE model, mainly in subtropical areas and due to changes in monsoon, but very similar to *Kaplan et al.* [2002], who show an increase of about 100 GtC for the last 8 ka.

The modelled carbonate ion concentration in the deep tropical ocean is shown in the supplemental material Fig. S1. The CO_3^{2-} concentration declines by 11 and 18 $\mu\text{mol}/\text{kg}$ in the Pacific and Atlantic basins, respectively. For the Atlantic, this is higher than the decrease by 10–12 $\mu\text{mol}/\text{kg}$ reconstructed by *Broecker et al.* [1999]. This mismatch could be caused by (i) imperfectness of the coarse zonally-averaged ocean model in CLIMBER2, (ii) complications of CO_3^{2-} reconstruction [*Barker and Elderfield*, 2002], (iii) overestimation of SWS in the model experiment, or (iv) neglect of the non-equilibrium carbonate chemistry state in the initial conditions.

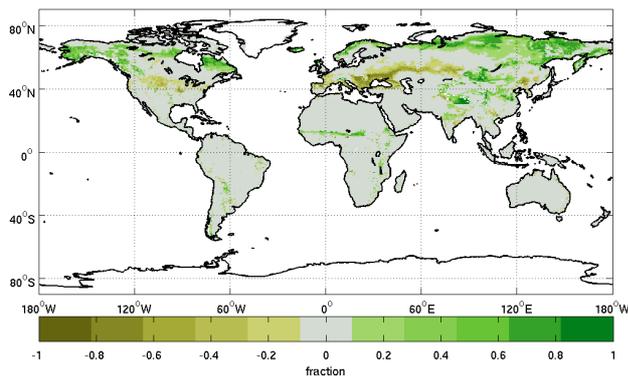


Figure 2. Tree cover change 8ka BP - present day in experiment AOV, expressed as grid cell fraction. Positive values imply higher tree cover at 8 ka BP than at present day.

Fig. 2 shows the difference in tree cover between 8 ka BP and present day (potential natural vegetation) for experiment AOV. A northward shift of the boreal forest margin at 8 ka BP is especially

pronounced in Siberia. These changes are in line with pollen-based reconstructions for western and central Siberia [*Bigelow et al.*, 2003]. In north-eastern North America, the simulated northward shift in forest margin, as compared to the present, is in disagreement with pollen data [*Williams*, 2003], but this can be expected since we neglected the remains of the Laurentide ice sheet. Tree cover was reduced between 30°N and 50°N, but vegetation cover at the northern Sahel border is expanded northward for 8 ka BP, although this shift is less pronounced than in the reconstructions [*Jolly et al.*, 1998].

4. Summary and conclusions

We have performed four model experiments in order to investigate the evolution of the global carbon cycle between 8 ka BP and present day. All our experiments without shallow water sedimentation of CaCO_3 show a decrease in atmospheric CO_2 between 8 ka BP and present day, with the decrease in CO_2 strongest in AOV where carbon uptake by peat is considered. *Frolking and Roulet* [2007] have assessed the climate forcing impact of peatlands by considering both CO_2 uptake and CH_4 release, and our assumption of sustained uptake of carbon by peat is in line with their conclusion of a net radiative cooling. Of our experiments, only AOVPC including SWS shows an increase in atmospheric CO_2 that is close to the reconstructions [*Monnin et al.*, 2004]. The model results suggest that the Holocene rise in CO_2 can be explained by a natural forcing, outgassing of CO_2 due to SWS. This is in line with the conclusions of *Ridgwell et al.* [2003], which we substantiate by using a coupled climate carbon cycle model.

When it comes to the $\delta^{13}\text{C}$ of CO_2 , our experiments are within the error bars of the latest measurement data [*Elsig et al.*, 2009], until anthropogenic deforestation can no longer be neglected at about 2.5 ka BP. While SWS in our experiment is at the higher end of estimates for the present day [*Milliman*, 1993; *Ridgwell et al.*, 2003], emissions may have been substantially higher between 8 ka BP and 4 ka BP, since sea level still rose by several meters between those times, likely leading to higher coral reef formation rates. At the same time, we are using a relatively low estimate of peat carbon uptake. While estimates vary, carbon uptake by peatlands may have been up to three times as large as considered here. This possible underestimation of the carbon uptake could easily be compensated if we assumed higher rates of CaCO_3 sedimentation or introduced a small contribution by carbonate compensation to the early Holocene vegetation regrowth, though the divergence in the $\delta^{13}\text{C}$ signal after 2.5 ka BP would increase at higher peat accumulation.

There are limitations to our study. In particular, we did not account for a reorganization of the oceanic circulation during the transition, which should have some imprint on the early Holocene CO_2 dynamics. We also neglected land carbon changes before 8 ka BP. Accounting for them will lead to some increase of CO_2 , as shown by *Elsig et al.* [2009]. However, glacial-interglacial changes in land carbon storage are not well constrained. While the *Elsig et al.* estimate of about 900 GtC of interglacial carbon buildup is within the envelope of uncertainties, recent studies provide evidence for enhanced glacial carbon storage in permafrost based on measurements [*Zimov et al.*, 2009] and isotopic analysis [*Ciais et al.*, 2009]. The latter study suggests less than 300 GtC transferred from ocean to land. Such a low transfer would invalidate carbonate compensation as the main explanation of the Holocene CO_2 growth. At the same time, the low carbon transfer is in line with the coral reef regrowth approach used in this paper. However, it is too early to rule out the biosphere regrowth hypothesis as observational constraints on it, as well as on the coral reef growth, are rather weak, and they both are in line with the decreasing CO_3^{2-} concentration in the deep ocean shown in sediment cores [*Broecker et al.*, 1999].

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